

DEPARTMENT OF GEOPHYSICS  
STANFORD UNIVERSITY  
STANFORD, CALIFORNIA 94305

FINAL REPORT

NASA GRANT NGL 05-020-232

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to the

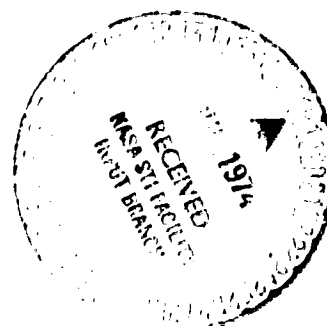
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## Introduction

This is a final report summarizing research accomplished under NASA grant NGL 05-020-232, a grant concerned with many aspects of lunar and planetary related research.

Because most of the scientific results have been published in the open literature we have only summarized these results in abstract form. The work accomplished under this grant during the last year and not yet published is summarized below.

## Seismic velocity studies pertinent to the lunar crust and mantle

Results from the lunar passive and active seismic experiments indicate that the moon has a 'crust' in which the velocity increases from a few hundred meters/second to a few km/s at a depth of 5 km or so, and a 'mantle' which is characterized by a very high velocity (7.7 to 9 km/sec). These results have suggested two interesting problems of the history and state of the moon: the problem of microcracks in the lunar crust, and the influence of temperature on seismic velocity in the lunar mantle.

It is generally accepted that the low seismic velocity in the lunar crust is caused by the high proportion of microcracks present in the rocks. Two theories can be invoked to explain the origin of these cracks:

a. The moon was formed by cold accretion from planetary dust, without extensive heating. The microcracks in the crust have not 'healed' since the accretion of the moon.

b. The moon passed through a 'hot' stage in its formation, during which its crust would have solidified except that meteor impacts have repeatedly shattered the crust.

Data from the Apollo 17 Lunar Seismic Profiling Experiment indicate that the depth of microcracking in the moon is actually surprisingly shallow. Cold accretion would imply a much deeper surficial low velocity layer. The lack of a thick surface zone implies either that the hypothesis of cold accretion is incorrect or that healing of microcracks, even at relatively low temperatures, is sufficient to eliminate them with time. Even faster 'healing' is required to explain the lack of deep microcracking due to meteor impacts. Simple scaling suggests that the shattering depth of an impact should be comparable with its diameter. This implies shattering to depths greater than several tens of kilometers. Thus, either our conception of the scaling law is erroneous, or else 'healing' is rapid enough to eliminate cracking with time.

Unlike the earth, pressure in the moon does not have a strong effect on the seismic velocity and consequently the state of the lunar mantle. Recent heat flow measurements on the moon imply, however, that the temperature in the moon's interior may be high.

We have developed a method to estimate the thermal conductivity of non-porous rocks and have applied it to systems of MgO-NiO, plagioclase, olivine, and pyroxene. It will be shown later that temperature may have a somewhat larger effect on the seismic velocity in the lunar mantle than previously believed.

The conclusion emerges that temperature, although not necessarily very high, plays an important role in the lunar history, in 'healing' cracks in the crust and determining seismic velocities in the mantle.

In the past several years under this grant we have carried out an extensive study of velocities in geological materials which are relevant to the lunar state.

a. Compressional and shear wave velocities in loose aggregates

The velocities of seismic compressional waves and, for the first time, shear wave velocities in silica sand, volcanic ash (Fig. 1) and basalt powder, were determined under hydrostatic confining pressures to 2.5 kilobars. Simultaneously, we have also obtained the porosity of these materials as a function of confining pressure. Although porosity undergoes very large permanent decreases due to applied compaction cycles and the extensive inelastic deformation, both  $V_p$  and  $V_s$  velocities of the compacted sample are equal to the velocities in the uncompressed powders. The velocities are a unique function of pressure and are independent of the pressure history of the sample. The derived effective elastic moduli and porosity and density are not unique functions of pressure. Poisson's ratio of the granular samples is abnormally high, even at two kilobars in comparison with solid rocks.

b. Effects of temperature on seismic velocity in compacted powders

The low seismic velocities in compacted powders suggested that it might be necessary to subject the powders to elevated temperatures in order to achieve equivalence with velocities in rocks. Samples of volcanic ash were cycled in high pressure and temperature experiments (Table 1). The initial high temperature experiments were conducted in a sophisticated, although awkward gas-medium apparatus. The observed increase in seismic velocity was sufficiently interesting that we decided to assemble a simpler, externally-heated, fluid-medium vessel. This apparatus is now operational, but studies of hot compaction in powders are as yet limited to only moderate temperatures.

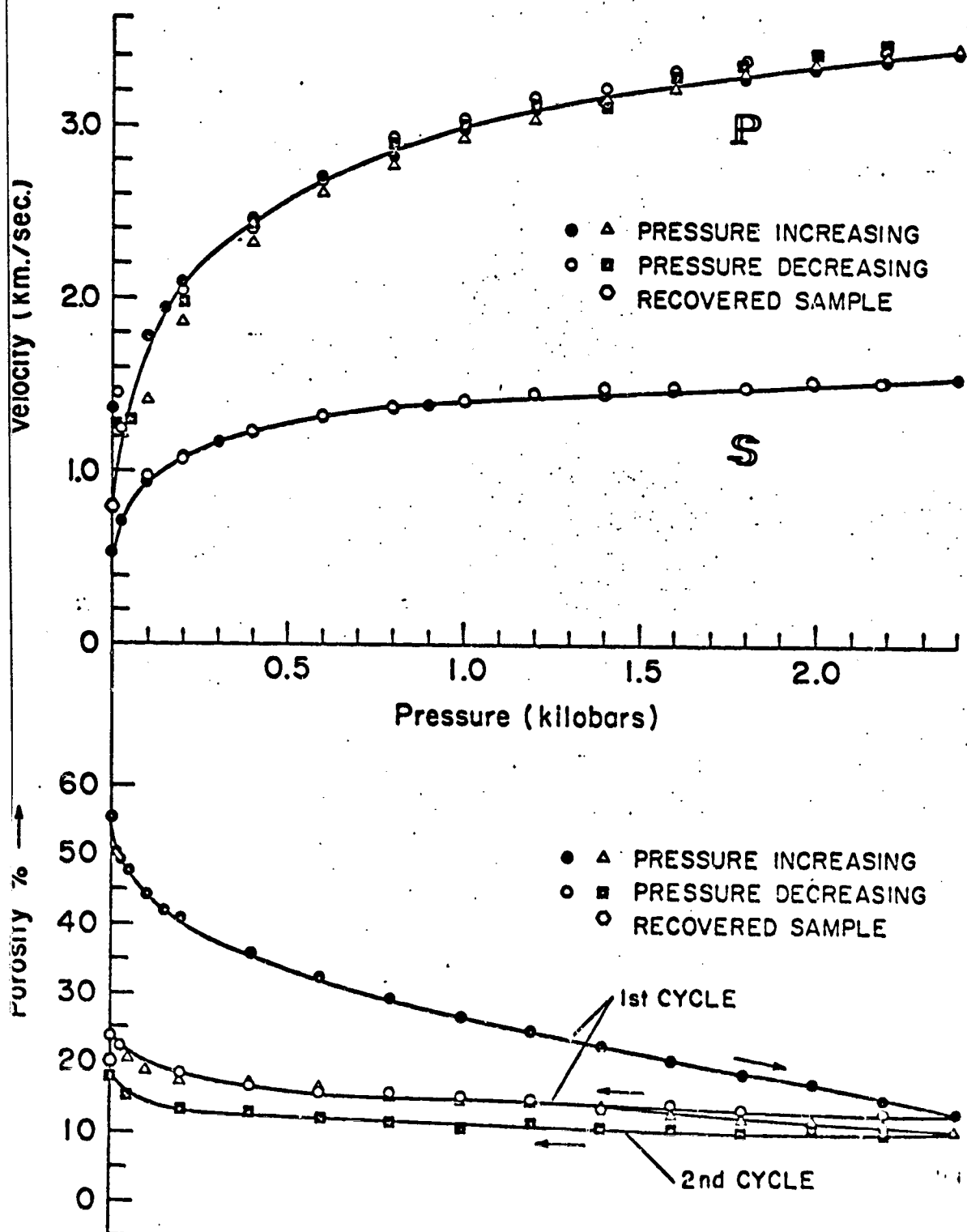


Fig. 1

TABLE 1

Compressional Velocity in Volcanic Ash Samples After Various P-T Cycling.  
Measurements at Room Conditions.

Pressure History	Temperature/time History	Compressional Velocity
2.5 kb	room temp.	.77 km/s
10.0 kb	room temp.	1.10 km/s
2 kb	-200°C/~1 hrs	1.90 km/s
2 kb	300°C/5 hrs	4.00 km/s

c. Effects of temperature and pressure on seismic velocities in crystalline rocks

We have measured the compressional and shear velocities of crystalline rocks to temperatures over 400°C, while the sample is under 5 kilobars hydrostatic confining pressure. In these experiments, the two velocities (Fig. 2) are measured concurrently as the sample (Westerly granite) is subjected to changes in temperature and pressure. Compressional velocities increase rapidly with increasing pressure below 2 kilobars, as the numerous microcracks in this sample are being closed. At higher pressures, in the more linear portion of the curve, the isotherms display the decrease in velocity as temperature increases. Temperature has a larger retarding effect on shear wave propagation, and the decrease in shear velocity is particularly large between 300 and 400°C.

The computed elastic moduli (Fig. 3) are somewhat lower than other published reports, and this is attributed to the microcracks which form during pressure-temperature cycling. The bulk moduli fall within the narrow shaded region, and there is no observable, systematic temperature dependence. In contrast, the shear modulus decreases markedly with



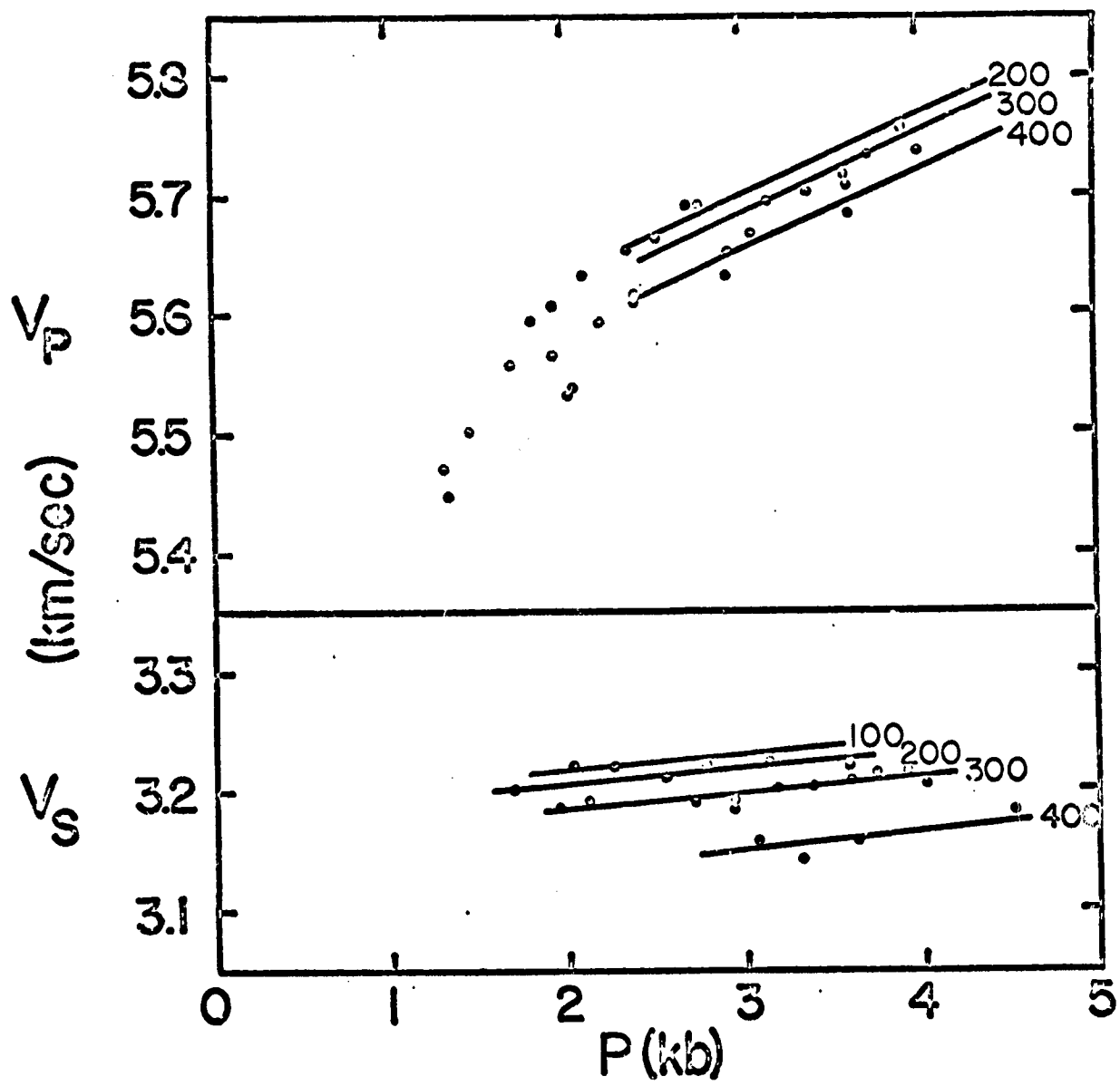


Fig. 2

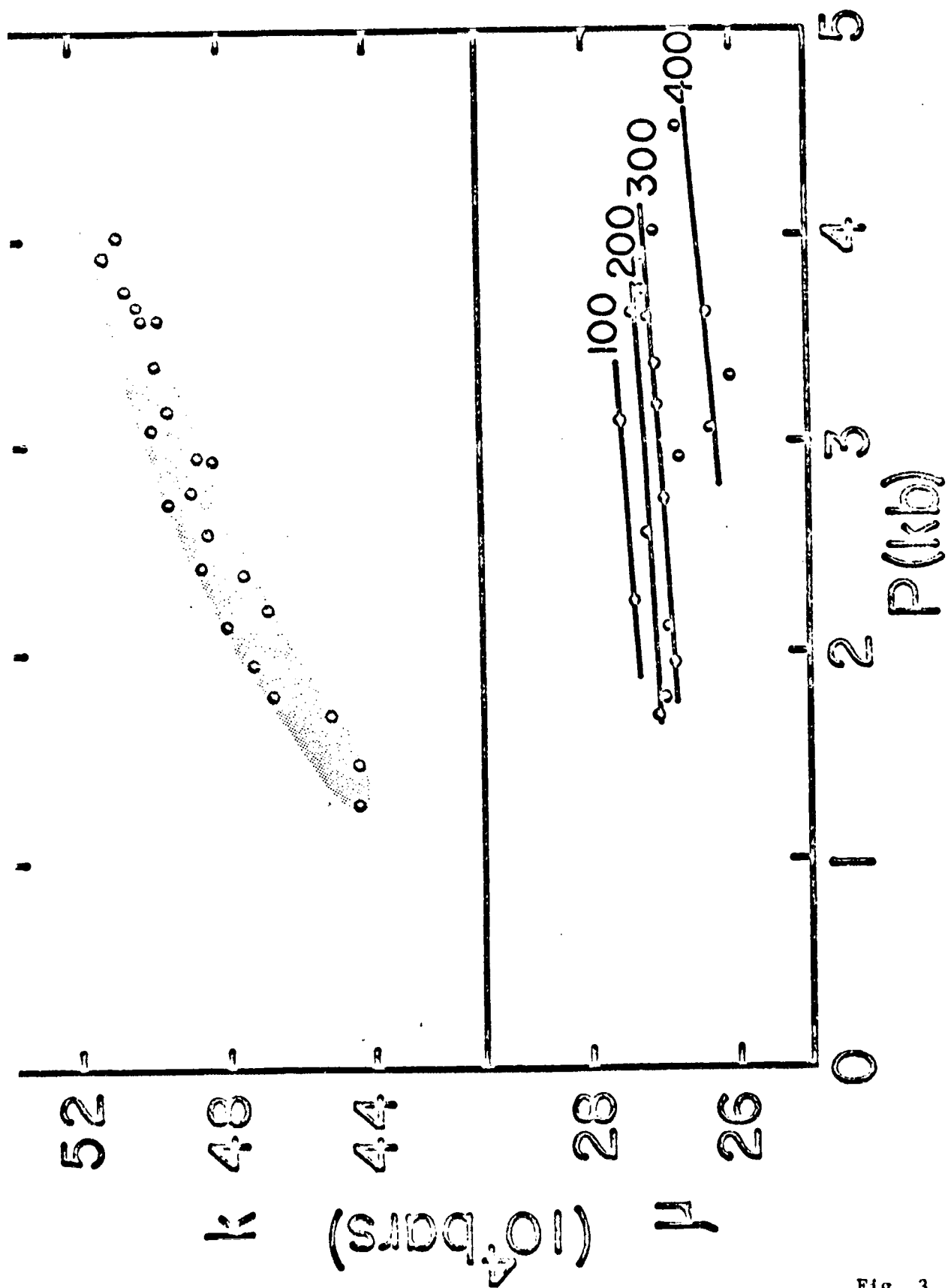


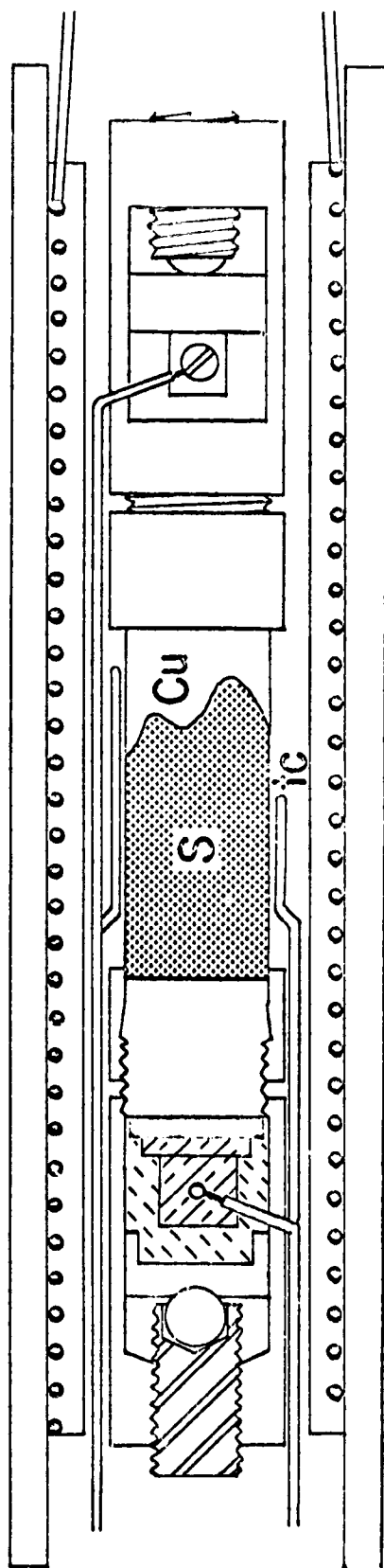
Fig. 3

increasing temperature. The decrease in compressional velocity with increasing temperature, is therefore largely due to a decrease in the shear modulus whereas the bulk modulus is relatively unaffected.

The concurrent measurement of compressional and shear velocities permits calculation of Poisson's ratio, and its value is of interest because one normally assumes that the "seismic ray parameter" is constant within the crust and upper mantle. The results show that both temperature and pressure act to increase Poisson's ratio. Pressure acts primarily by increasing the bulk modulus, while temperature markedly decreases the shear modulus of the aggregate.

Experiments are run in a large volume, internally-heated pressure vessel which attains argon pressures and temperatures to 8 kb and 800°C simultaneously. Figure 4 shows the sample assembly, the resistance furnace, and how the sample is sealed in an annealed copper jacket by deforming the copper between the steel end plug and the sleeve. The piezoelectric lithium niobate transducers are held mechanically against the steel end plugs on each end of the sample. After calibration for travel time through the steel end plugs, the velocities in the sample are determined from simple time-of-travel measurements.

We have measured compressional velocities above 800°C with Y-36 cut  $\text{LiNbO}_3$  transducers, but subsequent reduction of the transducer limits measurements to 550° after 48 hours. Alternatively we have used X-cut  $\text{LiNbO}_3$  transducers to generate clean shear arrivals and a small compressional forerunner, so that compressional and shear wave velocities can be measured simultaneously. Unfortunately, the piezoelectric constants of the crystal transducer change increasingly near 400°C so that the desired shear arrival soon cannot be isolated within the stronger compressional arrival.



5 cm

Fig. 4

In separate experiments of terrestrial interest, we have independently controlled the pore-fluid pressure inside the jacketed sample. We are examining whether the physical properties of super-critical water might produce 1.) seismic velocity inversions which have been tentatively identified in the earth's crust and 2) variations in  $V_p/V_s$  which are premonitory to some thrust-type earthquakes.

ABSTRACTS OF SCIENTIFIC PAPERS SUPPORTED

BY NASA GRANT NGL 05-020-232

Ahrens, T. J., C.F. Allen and R.L. Kovach, Explosive gas blast: the expansion of detonation products in vacuum, J. Applied Physics, 42, 815-829, 1970.

A series of 0.2- to 3-gm HNS charges were detonated in vacuums of  $10^{-3}$  to  $10^{-5}$  Torr. The resultant freely expanding, detonation product, gas blast achieves terminal velocities of 8 to 12 km/sec within 3 to 5  $\mu$ sec after the detonation wave arrives at the free surface. Measured pressure profiles display rise times to maximum stagnation ("reflected shock") pressure varying from  $\sim 30 \mu$ sec, 20 cm away from a 2.6 gm charge, to  $\sim 185 \mu$ sec, 127 cm away from 0.2 gm charge at  $10^{-5}$  Torr. Rise times were generally shorter at  $10^{-3}$  and  $10^{-4}$  Torr; the  $10^{-5}$  Torr values agree with numerical calculations. Using cube root scaling of charge mass, the observed peak reflected pressure as a function of range may be represented by

$$p = 6.5 \times 10^5 \text{ (bar)} r'^{-3.5},$$

where  $r'$  the ratio of the range to the equivalent charge radius.

Anderson, D.L. and R.L. Kovach, Universal dispersion tables III. free oscillation variational parameters, Bull. Seism. Soc. Amer., 59, 1667-1693, 1969.

The effect of a small change in any parameter of a realistic earth model on the periods of free oscillation is computed for both spheroidal and torsional modes. The normalized partial derivatives, or variational parameters, are given as a function of order number and depth in the earth. For a given mode it can immediately be seen which parameters and which regions of the earth are controlling the period of free oscillation. Except for  ${}_0S_0$  and its overtones the low-order free oscillations are relatively insensitive to properties of the core. The shear velocity of the mantle is the dominant parameter controlling the periods of free oscillation and density can be determined from free oscillation data only if the shear velocity is known very accurately. Once the velocity structure is well known free oscillation data can be used to modify the average density of the upper mantle. The mass and moment of inertia are then the main constraints on how the mass must be redistributed in the lower mantle and core.

Anderson, D.L. and R.L. Kovach, The lunar interior, Physics of Earth and Planetary Interiors, 6, 116-122, 1972.

The compressional velocities are estimated for materials thought to be important in the lunar interior and compared with lunar seismic results. The lower lunar crust has velocities appropriate for basalts or anorthosites. Anorthosite is preferred if lunar basalts result from a small degree of partial melting. The high velocities associated with the uppermost mantle imply high densities and a change to a lighter assemblage at depths of the order of 120 km. Ca- and Al-rich minerals are important components of both the lower crust and the upper mantle. Most of the moon may have accreted from refractory material rich in Ca, Al, U, and the rare-earth elements. The important mineral of the upper mantle is garnet; possible accessory minerals are kyanite, spinel and rutile. If the seismic results stand up, the high velocity layer in the moon is more likely to be a high-pressure form of anorthosite (garnet + kyanite + quartz) than eclogite, pyroxenite or dunite. The thickness of the layer is of the order of 50 km. The deep interior of the moon may also be rich in Ca and Al especially if temperatures are near solidus. Achondrites such as eucrites and howardites and some inclusions in Type III carbonaceous chondrites have more of the required characteristics of the lunar interior than carbonaceous chondrites. A density inversion in the moon is a strong possibility.

Booker, J.R., Mare Orientale gravity anomaly, Nature, 227, no. 5253, 56, 1970.

Sjogren has reported that there is an annular negative gravity anomaly surrounding a small positive anomaly over the central basin of Mare Orientale (paper given at the NATO Advanced Study Institute on the Moon and Planets, Newcastle, April 1970). I should like to point out that outside this negative ring, there is a positive ring and perhaps further alternating rings. This outer positive ring contains as much apparent surface mass as the inner negative ring. The integrated mass anomaly associated with the Orientale region is very nearly equal to the central positive anomaly.

Booker, J.R., Thermal state of the Moon, EOS, Transactions, American Geophysical Union, 51, 774, 1970.

I have investigated the temperature of the lunar interior using an approximate theory of finite amplitude convection in an internally heated, constant viscosity, self-gravitating sphere. This may be justified because the temperature effect on viscosity will be large only in a thermal boundary layer where the fluid velocity is small and heat conduction dominates. I am experimentally examining this hypothesis. For a moon with viscosity  $\eta = 700T \exp(46420/T)$  poise and heating  $q = 7.8 \times 10^{-3}$  ergs/gm sec, I find a thermal boundary layer thickness  $\epsilon = 245$  km and in the isothermal interior:  $T = 1150^\circ\text{K}$ ,  $\eta = 2.6 \times 10^{23}$  poise and Rayleigh number  $= 1.9 \times 10^5$ . The adiabatic gradient is not included. Increasing  $q$  by a factor of ten decreases  $T$  by  $50^\circ\text{K}$  and  $\epsilon$  by 17 km. Decreasing  $q$  has an opposite but nearly equal effect. The cell overturn time is less than one tenth of the moon's age justifying a steady state theory. Thus, the present lunar thermal state is probably dominated by the rheology rather than the heating rate or initial conditions.

Booker, J.R. and R.L. Kovach, Mascons and the moon's gravity field, Physics of Earth and Planetary Interiors, 4, 180, 1971.

The apparent surface mass  $M$  of the mascon in a lunar ringed mare is approximately proportional to the area  $A$  of the deep mare material in the basin:

$$M \propto A^{\sigma},$$

where  $\sigma$  lies between 0.85 and 1.20 with a most probable value of 1.02. This relationship can be interpreted in terms of thin plates of dense rock whose thickness is the same for all ringed maria.

Booker, J.R., R.L. Kovach and Lee Lu, Mascons and lunar gravity, J. Geophys. Research, 75, 6558-6564, 1970.

The mascon in a lunar ringed mare is approximately proportional to the area of the mare material in the basin. This relationship is consistent with the hypothesis that the lunar mascons are produced by dense plugs in the maria, and it means that the maximum thickness of the uncompensated rock is the same for all maria. The relationship also predicts the presence of mascons in other ringed lunar structures, such as Maria Orientale and Smythii, which are consistent with satellite Doppler data. The relative masses of the known and predicted mascons accurately predict the moon's dynamical asymmetry without any large mascons on the lunar farside. However, reconciliation of the absolute differences between the lunar moments of inertia with satellite accelerations directly above the maria requires mascons buried deeper than 250 km. Such deeply buried mascons seem unlikely. It therefore also seems unlikely that the differences in the lunar moments of inertia are completely due to the mascons. However, a converse relationship cannot be ruled out. Examination of the degree variances of harmonic analyses of lunar gravity reveals a gentle peak near degree 10. This peak is predicted by the spacing of the two largest mascons, Imbrium and Serenitatis.

Cooper, M.R., R.L. Kovach and J.S. Watkins, Lunar near-surface structure, Reviews of Geophysics and Space Physics, 12, August 1974.

Seismic refraction data obtained at the Apollo 14, 16, and 17 landing sites permit a compressional wave velocity profile of the lunar near-surface to be derived. Although the regolith is locally variable in thickness, it possesses surprisingly similar seismic characteristics. Beneath the regolith at the Apollo 14 Fra Mauro site and the Apollo 16 Descartes site is material with a seismic velocity of ~300 m/s, believed to be brecciated material or impact derived debris. Considerable detail is known about the velocity structure at the Apollo 17 Taurus-Littrow site. Seismic velocities of 100, 327, 495, 960, and 4700 m/s are observed. The depth to the top of the 4700 m/s material is 1385 m, compatible with



gravity estimates for the thickness of mare basaltic flows, which fill the Taurus-Littrow valley. The observed magnitude of the velocity change with depth and the implied steep velocity-depth gradient of  $\sim 2$  km/s/km are much larger than have been observed on compaction experiments on granular materials and preclude simple cold compaction of a fine-grained rock powder to thicknesses of the order of kilometers. The large velocity change from 960 to 4700 m/s is more indicative of a compositional change than a change of physical properties alone. This high velocity is believed to be representative of the material that forms the lunar highlands.

Fujii, N., A method to estimate effective thermal conductivity of non-porous rocks, EOS, Transactions, American Geophysical Union, 55, 422, 1974.

The essential factors controlling thermal conductivity of a non-porous, macroscopically isotropic and homogeneous rock, such as igneous rocks, are volume fractions and thermal conductivities of constituent minerals. Thermal conductivity of minerals which form a solid solution shows wide variation with composition and has a minimum value at an intermediate composition. To explain the variation of thermal conductivity in a series of solid solution, the mechanism is assumed to be an imperfection scattering by the mass difference between mutually soluble ions. A simplified formula is derived and applied to the systems of MgO-NiO, plagioclase, olivine and pyroxene. To obtain effective thermal conductivity of rocks a 'perfectly disordered' distribution is assumed for the spatial distribution of grains of constituent minerals. This is essentially the same as that previously proposed for elastic properties. Using available data for minerals, calculated values by this method are compared with that directly measured for several igneous rocks. The relative differences between these are almost within 10% for granites, gabbro, basalts, andesite, and diorite, though the data might include some uncertainty.

Kovach, R.L., The velocity structure of the moon, in Advanced Space Experiments, Advances in the Astronautical Sciences, 25, 391-400, 1969.

The question of the distribution of seismic velocities in the Moon is examined in the light of current data. Thermal history calculations combined with laboratory measurements on rocks lead to the conclusion that a broad low-velocity zone exists in the Moon together with possible extensive melting at depth. It will be difficult to decipher lunar seismic data from only a single sensor; an array of widely spaced, simultaneously recording sensors is required. Some aspects of explosion seismology on the Moon are discussed.

Kovach, R.L. and T.J. Ahrens, Explosive seismic sources for the moon, Geophysics, 35, 33-44, 1970.

The coupling of seismic energy under vacuum conditions, such as the moon, using an untamped surface charge is different from coupling in air. In vacuum, the explosive gas blast and the detonation products continuously expand outward and interact with the solid surface.

A series of model experiments was performed to investigate the effect of vacuum on coupling seismic energy. HNS charges of 0.2 gm each were detonated in contact with a plate and block of acrylic plastic in vacuum and in air. The amplitudes of the first and second arrivals (longitudinal and shear plate wave) are about 50 percent greater in vacuum than in air because the plate velocity ( $\sim 2.4$  km/sec) more closely matches the gas-blast velocity ( $\sim 3.5$  to  $7.5$  km/sec) than the sound-wave velocity ( $\sim 0.35$  km/sec). When the charges are detonated in contact with the block to generate direct body waves, little difference is noted in the first arrival amplitudes in air and vacuum; suspending the charge one charge-diameter above the surface produces about 25 percent lower first amplitudes in a vacuum. Large scale experiments were also performed in air to examine the effect of the detonation on seismic coupling.

Kovach, R.L. and R. Robinson, Upper mantle structure in the Basin and Range province, western North America, from the apparent velocities of S waves, Bull. Seism. Soc. Amer., 59, 1653-1665, 1969.

The variation of shear velocity with depth in the upper mantle for the Basin and Range province of western North America has been studied with direct measurements of  $dT/d\Delta$  for S waves in the distance range  $14^\circ < \Delta < 40^\circ$ . Three orthogonal components of digital data were used and onset times were determined using the product of the horizontal radial and vertical components of motion and particle motion diagrams. A linear LRSM array in Arizona was used for the measurement of  $dT/d\Delta$ .

An S-wave velocity distribution is derived, compatible with P-wave velocity models for the same region. The derived model consists of a thin lid zone of shear velocity  $4.5$  km/sec overlying a low-velocity zone and a change in velocity gradient at a depth of  $160$  km. Two regions of high-velocity gradient are located at depths beginning at  $360$  km and  $620$  km.

Kovach, R.L. and J.S. Watkins, The velocity structure of the lunar crust, The Moon, 7, 63-75, 1973.

Seismic refraction data, obtained at the Apollo 14 and 16 sites, when combined with other lunar seismic data, allow a compressional wave velocity profile of the lunar near-surface and crust to be derived. The regolith, although variable in thickness over the lunar surface, possesses surprisingly similar seismic properties. Underlying the regolith at both the Apollo 14 Fra Mauro site and the Apollo 16 Descartes site is low-velocity brecciated material or impact derived debris. Key features of the lunar seismic velocity profile are:

(i) velocity increases from  $100\text{--}300$  ms<sup>-1</sup> in the upper  $100$  m to  $\sim 4$  km s<sup>-1</sup> at  $5$  km depth, (ii) a more gradual increase from  $\sim 4$  km s<sup>-1</sup> to  $\sim 6$  km s<sup>-1</sup> at  $25$  km depth, (iii) a discontinuity at a depth of  $25$  km and (iv) a constant value of  $\sim 7$  km s<sup>-1</sup> at depths from  $25$  km to about  $60$  km.

The exact details of the velocity variation in the upper  $5$  to  $10$  km of the moon cannot yet be resolved but self-compression of rock powders

cannot duplicate the observed magnitude of the velocity change and the steep velocity-depth gradient. Other textural or compositional changes must be important in the upper 5 km of the moon. The only serious candidates for the lower lunar crust are anorthositic or gabbroic rocks.

Kovach, R.L. and J.S. Watkins, The structure of the lunar crust at the Apollo 17 site, Proc. Fourth Lunar Sci. Conf., Geochim. Cosmochim. Acta, Suppl. 4, vol. 3, 2549-2560, 1973.

A seismic profiling experiment was successfully executed on the lunar surface during the Apollo 17 mission allowing a determination of the structure of the lunar crust to a depth of several kilometers. The most outstanding feature of the seismic velocity variation, in the Taurus-Littrow region, is the stepwise increase with depth. A total thickness of about 1200 meters for the infilling mare basalts at the 17 landing site is also indicated from the seismic results. The apparent velocity is high (about 4 km/sec for P waves) in the material below the basalts.

Lu, L., J.R. Booker and R. L. Kovach, Mascons and the farside of the moon, EOS, Transactions, American Geophysical Union, 50, 638, 1969.

Satellite gravity observations of the moon have been used to infer the presence of lunar mascons near the limbs and on the lunar farside. An examination of the RMS power in the spherical harmonic coefficients of the moon's gravitational potential reveals peaks near degree 0 and 12. Such a bimodal distribution is expected if the moon's mass were homogeneous except for the presence of small superimposed mascons. However, construction of a model of the moon's gravity field is complicated by the lack of direct farside observations. Errors in orbit determination produce a much larger effect on estimates of the existence and position of mascons on the farside than on the nearside. We are examining the validity of lunar backside mascons which are consistent with observational errors. Instead of utilizing a field model which best satisfies the orbital observations in a least squares sense we are developing models which only satisfy the data to within the known observational error but which, in addition, satisfy smoothness criteria on the farside.

McAllister, B.D., J. Kerr, J. Zimmer, R.L. Kovach and J. Watkins, A seismic refraction system for lunar use, IEEE Transactions of Geoscience Electronics, GE-7, 164-171, 1969.

A miniaturized seismic refraction system has been constructed for possible use on early manned lunar landings. The detection system consists of three geophones, a three-channel amplifier, a geophone calibrator, and a logarithmic compressor system. A grenade launching device and an astronaut-held thumper staff are used as the sources of seismic energy. Seismic energy from the explosive sources is de-

tected, amplified, logarithmically compressed, converted to digital form, and formatted for real-time transmission to earth. Total weight of the seismic detection system and ancillary electronics, exclusive of the explosive sources, is 6.25 lb.

Robinson, R. and R.L. Kovach, Shear wave velocities in the earth's mantle, Physics of Earth and Planetary Interiors, 5, 30-44, 1972.

Direct measurements of the travel time gradient  $dT/d\Delta$  for S waves over the distance range  $14^\circ < \Delta < 100^\circ$ , together with travel time data, are used to derive a shear velocity model for the earth's mantle. A network of seismograph stations in Arizona operated as an array was used for the measurement of  $dT/d\Delta$ . The complex velocity structure in the upper mantle makes the use of multiple arrivals necessary to define the  $dT/d\Delta$ - $\Delta$  curve for distances less than  $55^\circ$ . In order to satisfy the data it is necessary to discard the usual assumption of lateral homogeneity below shallow depths and a shear velocity differential of up to 0.1 km/s down to a depth of 1000 km is proposed between western North America and areas of the Pacific Ocean. Distinctive features of the velocity model for the upper mantle beneath western North America are a low-velocity zone centered at 100 km depth and zones of high velocity gradient beginning at 400, 650 and 900 km. Also, the moderate velocity gradient derived for the depth range 150-400 km implies a density reversal in the same area if Love wave dispersion data are to be satisfied. The velocity model derived for the lower mantle beneath oceanic areas has a zone of high velocity gradient at 1200 km depth and a pronounced low-velocity zone at the core-mantle boundary.

Spencer, J.W. and A. Nur, Time-dependent velocity of water-saturated granite at high temperatures and pressures, EOS, Transactions, American Geophysical Union, 54, 1209, 1973.

The compressional velocity of a saturated sample of Westerly granite, enclosed in a leak-proof flexible jacket, is measured as the sample is subjected to repeated cyclic changes in pressure and temperature. Velocity changes over the range of  $P = 1.1 - 2.3$  kb and  $T = 23 - 46^\circ\text{C}$  are reversible with  $P$  and  $T$  when the sample is maintained at  $460^\circ\text{C}$  for 17 hrs,  $V_p$  increases by 1.0% during the waiting period, and when maintained at  $460^\circ\text{C}$  for 47 hrs,  $V_p$  increases by 1.5%. Conversely, maintaining the sample at  $23^\circ\text{C}$  and 1.1 kb for a sufficient time returns  $V_p$  to its original value. Two higher-pressure cycles show 2.0% velocity increases after 16 hrs at  $510^\circ\text{C}$  and 5.5 kb and 22 hrs at  $495^\circ\text{C}$ . The change during an additional 24 hrs is small. Time constants for these sluggish velocity changes appear too slow for pore fluid flow in this sample because of the high permeability at low effective pressure. The magnitudes of the velocity increase and the observed  $P$ - $T$  dependence of the time constants can be reasonably explained by solution of quartz in supercritical water (0.4 wt.% at  $460^\circ\text{C}$  and 2.3 kb and 0.8 wt.% at

510°C and 5.5 kb, Anderson and Burnham, 1965) which increases the viscosity and shear modulus of the pore fluid, and hence  $V_p$ .

Talwani, P., A. Nur and R.L. Kovach, Compressional and shear wave velocities in granular materials to 2.5 kilobars, J. Geophys. Research, **78**, 6899-6909, 1973.

In spite of the common occurrence of granular materials, both on the earth and on the moon, little is known about their mechanical behavior at pressures corresponding to large depths of burial. This is particularly true for seismic wave velocities. We have measured the compressional and shear wave velocities in silica sand, volcanic ash, and powdered basalt subjected to hydrostatic pressures from 1 atm to 2.5 kb. Simultaneous determination of porosity was also obtained as a function of confining pressure. The results reveal that large velocity (both  $V_p$  and  $V_s$ ) variations are almost completely reversible upon pressure cycling, even though there is an irreversible change in porosity. The amount of permanent change in porosity was found to depend on the original grain size of the sample. The derived effective elastic moduli are not unique functions of pressure. The change in  $V_p$  and the computed bulk modulus depends on the initial porosity, whereas grain size controls the behavior of  $V_s$  and the shear modulus. The rock powders have very large values for Poisson's ratio (~0.3-0.4) at high pressures, and it is thus verified that simple cold compaction to a pressure of a few kilobars cannot transform rock powders to solid rocks. These results have important implications for the self-compaction hypothesis, which has been postulated to explain the lunar near-surface seismic velocity variation.

Talwani, P., A. Nur and R.L. Kovach, Implications of elastic wave velocities for Apollo 17 rock powders, Proc. Fifth Lunar Sci. Conf., Geochim. Cosmochim. Acta, 1974.

Ultrasonic P- and S- wave velocities of lunar rock powders 172701, 172161, 170051 and 175081 were measured at room temperature and to 2.5 kb confining pressure. The results compare well with those of terrestrial volcanic ash and powdered basalt. P- wave velocity values up to pressures corresponding to a lunar depth of 1.4 km preclude cold-compaction alone as an explanation for the observed seismic velocity structure at the Apollo 17 site. Application of small amounts of heat with simultaneous application of pressure causes rock powders to achieve equivalence of seismic velocities for competent rocks.

Talwani, P. and A. Nur, Experimental determination of static and dynamic compressibility of rock powders, EOS, Transactions, American Geophysical Union, **55**, 415, 1974.

Experimental determination of  $V_p$ ,  $V_s$ , and porosity ( $\phi$ ) and bulk density of rock powders subjected to hydrostatic pressures up to 2.4 kb (Talwani et al. 1973) enabled us to compute both the static and dynamic compressi-

bility ( $\beta_s$  and  $\beta_d$ ).  $\beta_s$  was found to depend on the rate of change of porosity with pressure  $d\phi/dp$ , as had been theoretically predicted by Walsh (1965) while  $\beta_d$  had a  $\frac{1}{\phi} \frac{d\phi}{dp}$  dependence. Both  $\beta_s$  and  $\beta_d$  varied linearly with respect to each other.  $\beta_s$  is  $> \beta_d$  at low confining pressures.

Wesson, R. L., A time integration method for computation of the intensities of seismic rays, Bull. Seism. Soc. Amer., 60, 307-316, 1970.

A finite-difference time integration method for the calculation of seismic ray intensity is developed. Discontinuities in the depth derivatives of the velocity-depth function at layer boundaries cause anomalies in the intensity distance curves calculated using the standard integral formulation. The time integration method overcomes these difficulties. Calculations for a simple analytic case and a Gutenberg earth model demonstrate the difficulties with the standard integral method and the superior performance of the time integration scheme. The method may also be applied to laterally inhomogeneous earth models.

Wesson, R.L., Travel-time inversion for laterally inhomogeneous crustal velocity models, Bull. Seism. Soc. Amer., 61, 729-746, 1971.

The realistic interpretation of seismic travel-time data from structurally complex areas, and the accurate location of earthquake hypocenters in such areas, require seismic ray computations for laterally inhomogeneous velocity models. Numerical simulation of the ray differential equations provides a practical means of performing the necessary calculations. A least-squares scheme is used to obtain models which fit travel-time data and are consistent with geological data. Laterally inhomogeneous velocity models are obtained for travel-time data from explosions for two areas in California: the Bear Valley area, 40 km southeast of Hollister and the Borrego Mountain area, 160 km northeast of San Diego. Both regions are characterized by a substantial lateral variation of seismic velocity, and the derived models exhibit most of the significant structural features of the areas. An algorithm for the direct solution of ray boundary value problems, based on the iterative solution of a tridiagonal set of simultaneous equations, allows for the input of geophysical intuition in finding the rays between a source and a station.